CORRESPONDENCE

Comments on “Langmuir Turbulence and Surface Heating in the Ocean Surface Boundary Layer”

YIGN NOH AND YEONJU CHOI
Department of Atmospheric Sciences, Yonsei University, Seoul, South Korea

(Manuscript received 4 July 2017, in final form 22 January 2018)

ABSTRACT

Using large-eddy simulations (LES) it is shown that the depth of a diurnal thermocline \(h\) should be scaled by the Zilitinkevich scale \(L_Z\), not by the Monin–Obukhov length scale \(L_{MO}\), contrary to the proposition by Pearson et al. Their argument to explain the slower increase of \(h\) than \(L_{MO}\) using the effect of the preexisting thermocline is also invalid.

In a recent paper, Pearson et al. (2015, hereinafter PGPB) suggested, using LES, that the depth of a diurnal thermocline \(h\) can be estimated by

\[
h = \beta L_{MO}/[\beta L_{MO}/(h_0 + 1)],
\]

where \(L_{MO}\) is the Monin–Obukhov scale \((= u_*/Q_0)\), \(u_*\) is the frictional velocity, \(Q_0\) is the surface buoyancy flux, \(h_0\) is the depth of a preexisting thermocline, and \(\beta\) is a proportionality constant. PGPB considered \(L_L (= w^3_{*L}/Q_0)\), instead of \(L_{MO}\), where \(w^3_{*L} = (u^3_*/U_S)^{1/3}\) and \(U_S\) is the Stokes drift velocity at the surface, but \(L_L\) is proportional to \(L_{MO}\) if \(L_L = L_{MO}\) is constant \((= 0.3)\), as in their simulations, since \(L_L = \text{La}^{-2} L_{MO}\).

The prediction that the depth of a diurnal or seasonal thermocline is proportional to \(L_{MO}\) was originally proposed by Kraus and Turner (1967) from the balance of the contributions from wind stress \(u^3_*\) and surface heating \(Q_0h\) in the TKE budget of the mixed layer. However, Goh and Noh (2013) showed recently, using LES, that the Coriolis force plays a critical role in the formation of a seasonal thermocline, and \(h\) should be scaled by the Zilitinkevich scale \(L_Z \equiv u^3_*/[fQ_0]^{1/2}; Zilitinkevich 1972\) instead of \(L_{MO}\), where \(f\) is the Coriolis parameter. It implies that the downward transport of momentum is limited to the Ekman boundary layer \(\lambda (= u_*/f)\), and therefore the contribution from wind stress should be modified to \(u^3_*/(\lambda/h)\), which leads to \(h \propto L_Z\) from the balance with \(Q_0h\) (Goh and Noh 2013). The new scaling has also been verified from the analyses of observation data (Yoshikawa 2015; Lee et al. 2015). Since diurnal and seasonal thermocline formations share a common feature, that is, the interaction between downward turbulent mixing of heat and the suppression of turbulence by stratification (Noh 1996; Noh et al. 2009; Goh and Noh 2013), it is natural to expect that the depth of a diurnal thermocline \(h\) is also scaled by \(L_Z\), rather than \(L_{MO}\).

Therefore, we carried out LES under conditions similar to PGPB in order to examine the scaling of \(h\). The LES model is the same as that used in Noh et al. (2004, 2009) and Goh and Noh (2013), in which both Langmuir circulation and wave breaking are realized. The wavelength and height for \(U_S\) were fixed as 40 and 1 m, and \(u_*\) was fixed as \(u_* = 0.01 \text{ m s}^{-1}\), resulting in \(\text{La} = 0.45\).

The simulation was carried out to reproduce the formation of a diurnal thermocline under constant \(u_*\) and \(Q_0\), starting either from the homogeneous mixed layer with uniform density and from the stratified layer with a preexisting thermocline. Integration was carried out initially without surface heating for 12 h to develop a neutral boundary layer and then continued until \(h\) reaches an equilibrium after the onset of surface heating. The time of the onset of surface heating is set to be \(t = 0\) h, which corresponds to the time of sunrise. The

Corresponding author: Yign Noh, noh@yonsei.ac.kr

DOI: 10.1175/JPO-D-17-0135.1
© 2018 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the AMS Copyright Policy (www.ametsoc.org/PUBSReuseLicenses).
depth of a preexisting thermocline $h_0$ is defined by the depth of a thermocline at the onset of surface heating ($t = 0 \text{ h}$) after the mixed layer deepens by wind mixing for 12 h starting from the initial depth of 0, 10, 20, and 30 m. Simulations were performed with various different surface buoyancy fluxes ($Q_0 = 1.25, 2.5, 5 \times 10^{-7} \text{ m}^2\text{s}^{-3}$), Coriolis forces ($f = 0.25, 0.5, 1, \text{ and } 1.4 \times 10^{-4} \text{ s}^{-1}$), and depths of a preexisting thermocline $h_0$ and stratifications below it ($N^2 = 1 \text{ and } 5 \times 10^{-4} \text{ s}^{-2}$). The model domain is 300 m horizontally and 80 m vertically, and the grid size is 1.25 m in all directions. The physical nature of diurnal thermocline formation, simulated by LES, is explained in detail in Noh et al. (2009).

Figure 1 compares the evolutions of mean buoyancy profiles $B(z)$ after the start of surface heating ($Q_0 = 5 \times 10^{-7} \text{ m}^2\text{s}^{-3}$) with different $f$ ($f = 0.25 \times 10^{-4} \text{ s}^{-1}, 1.4 \times 10^{-4} \text{ s}^{-1}$) for both cases from the homogeneous layer and from a preexisting thermocline. Here, horizontal lines represent the depth of thermocline formation $h$ (red) from a preexisting thermocline with the depth $h_0$.
The depth of thermocline formation $h$ from the homogeneous layer, shown in Figs. 1a and 1c, is also included in Figs. 1b and 1d for comparison, and it will be called $h^*$ hereinafter. In the present work, both $h$ and $h_0$ are calculated by the depth of the maximum $N^2 (=\partial B/\partial z)$, as most widely used. On the other hand, PGPB calculated $h$ and $h_0$ through different methods: the depth where the linearly fitted line of buoyancy flux near the surface goes to zero for $h$ and the depth of a neutrally stratified layer for $h_0$. Hereinafter, $h$ obtained by the present method and by PGPB will be called $h_N$ and $h_{wb}$, respectively. Here, $h$ and $h^*$ are overlapped for $h_N$ in Fig. 1b and very close for $h_{wb}$ in Figs. 1b and 1d. The cases with $f = 0.25 \times 10^{-4} \text{ s}^{-1}$ do not reach equilibrium yet. It clearly shows that the formation of a diurnal thermocline, and thus $h$, is strongly affected by $f$, in contradiction to (1).

Variations of $h$ from the homogeneous layer, or $h^*$, with the length scales clearly illustrate that $h$ should be scaled by $L_Z$, instead of $L_{MO}$, for both $h_N$ and $h_{wb}$, as in the case of a seasonal thermocline (Goh and Noh 2013; Fig. 2). The proportional constant $\alpha = 0.7$ in the relation $h = \alpha L_Z$ is obtained from the data with $f \approx 0.5 \times 10^{-4} \text{ s}^{-1}$, in which the diurnal thermocline depth under equilibrium can be clearly identified (Fig. 2b). It is somewhat larger than in the seasonal thermocline ($\alpha = 0.5$; Goh and Noh 2013) in which $Q_0$ varies along the diurnal cycle.

Most data in Fig. 9a of PGPB were obtained with constant $f = 10^{-4} \text{ s}^{-1}$, which helps make the scatter in their figure smaller. Nonetheless, their own figure still shows clearly the decrease of $h$ with $f$. For example, at $L_{MO} = 33 \text{ m}$ ($L_L = 370 \text{ m}$) in their figure, $h = 32 \text{ m}$, when $f = 1.4 \times 10^{-4} \text{ s}^{-1}$, but $h = 45 \text{ m}$, when $f = 0.5 \times 10^{-4} \text{ s}^{-1}$, which implies that $h$ decreases with $f^{1/2}$ (see also Fig. 2a). They could not also include the data from $f = 0 \text{ s}^{-1}$, although the scaling by $L_{MO}$ does not exclude this case.

Another problem in Fig. 9a in PGPB is that the increasing rate of $h$ is slower than $h \propto L_{MO}$, which is the prediction by Kraus and Turner (1967; see also Fig. 2a). Considering that $u_0$ and $f$ are fixed during their simulations, their scatterplot actually represents the relation $h \propto L_{MO}^{1/2} \propto Q_0^{-1/2}$, which comes from $h \propto L_Z \propto Q_0^{-1/2}$. However, PGPB attempted to explain the reason for the slower increase than $h \propto L_{MO}$ through the influence of a preexisting thermocline at $z = h_0$.

It is clear from Fig. 1b, however, that $h$ is not affected by $h_0$, that is, $h = h^*$, as long as $h$ (or $h^*$) is sufficiently smaller than $h_0$ ($h/h_0 \approx 0.5$ for $h_N$ in Fig. 1b), contrary to (1). Figure 10a of PGPB shows that $h$ is much smaller than $h_0$ in most cases; that is, $h/h_0 \ll 1$ for $h_{wb}$. Figure 1b indicates that it corresponds to even smaller $h/h_0$ for $h_N$. It is therefore highly unlikely that $h$ is influenced by $h_0$ in their simulations. Note that the turbulence structure within the mixed layer with the depth $h < h_0$ may not remember $h_0$ long after decoupling from $h_0$ ($t \gg h/u_0$). Unfortunately, they did not perform the simulation from the homogeneous layer, corresponding to $h_0 = \infty \text{ m}$ in

![Fig. 2. Variation of $h$ with the length scales in the case from the homogeneous layer. Solid and blank symbols represent $h_N$ and $h_{wb}$, respectively. Here, $h$ and $h^*$ are overlapped for $h_N$ in Fig. 1b and very close for $h_{wb}$ in Figs. 1b and 1d. The cases with $f = 0.25 \times 10^{-4} \text{ s}^{-1}$ do not reach equilibrium yet. It clearly shows that the formation of a diurnal thermocline, and thus $h$, is strongly affected by $f$, in contradiction to (1). Variations of $h$ from the homogeneous layer, or $h^*$, with the length scales clearly illustrate that $h$ should be scaled by $L_Z$, instead of $L_{MO}$, for both $h_N$ and $h_{wb}$, as in the case of a seasonal thermocline (Goh and Noh 2013; Fig. 2). The proportional constant $\alpha = 0.7$ in the relation $h = \alpha L_Z$ is obtained from the data with $f \approx 0.5 \times 10^{-4} \text{ s}^{-1}$, in which the diurnal thermocline depth under equilibrium can be clearly identified (Fig. 2b). It is somewhat larger than in the seasonal thermocline ($\alpha = 0.5$; Goh and Noh 2013) in which $Q_0$ varies along the diurnal cycle. Most data in Fig. 9a of PGPB were obtained with constant $f = 10^{-4} \text{ s}^{-1}$, which helps make the scatter in their figure smaller. Nonetheless, their own figure still shows clearly the decrease of $h$ with $f$. For example, at $L_{MO} = 33 \text{ m}$ ($L_L = 370 \text{ m}$) in their figure, $h = 32 \text{ m}$, when $f = 1.4 \times 10^{-4} \text{ s}^{-1}$, but $h = 45 \text{ m}$, when $f = 0.5 \times 10^{-4} \text{ s}^{-1}$, which implies that $h$ decreases with $f^{1/2}$ (see also Fig. 2a). They could not also include the data from $f = 0 \text{ s}^{-1}$, although the scaling by $L_{MO}$ does not exclude this case. Another problem in Fig. 9a in PGPB is that the increasing rate of $h$ is slower than $h \propto L_{MO}$, which is the prediction by Kraus and Turner (1967; see also Fig. 2a). Considering that $u_0$ and $f$ are fixed during their simulations, their scatterplot actually represents the relation $h \propto L_{MO}^{1/2} \propto Q_0^{-1/2}$, which comes from $h \propto L_Z \propto Q_0^{-1/2}$. However, PGPB attempted to explain the reason for the slower increase than $h \propto L_{MO}$ through the influence of a preexisting thermocline at $z = h_0$. It is clear from Fig. 1b, however, that $h$ is not affected by $h_0$, that is, $h = h^*$, as long as $h$ (or $h^*$) is sufficiently smaller than $h_0$ ($h/h_0 \approx 0.5$ for $h_N$ in Fig. 1b), contrary to (1). Figure 10a of PGPB shows that $h$ is much smaller than $h_0$ in most cases; that is, $h/h_0 \ll 1$ for $h_{wb}$. Figure 1b indicates that it corresponds to even smaller $h/h_0$ for $h_N$. It is therefore highly unlikely that $h$ is influenced by $h_0$ in their simulations. Note that the turbulence structure within the mixed layer with the depth $h < h_0$ may not remember $h_0$ long after decoupling from $h_0$ ($t \gg h/u_0$). Unfortunately, they did not perform the simulation from the homogeneous layer, corresponding to $h_0 = \infty \text{ m}$ in...
It is unlikely to recover the relation $h \propto L_{MO}$ from this simulation, as (1) predicts. Finally, Fig. 3 shows how $h/h_0$ varies with $L_Z/h_0$. Indeed, it shows that $h$ is not affected by $h_0$ as long as $h_0$ is sufficiently larger than $h^*$ ($L_Z/h_0 < 0.9$). On the other hand, as $L_Z/h_0$ increases, $h$ becomes affected by $h_0$; $h$ can become even larger than $h_0$, as shown in Fig. 1d. Its increase is ultimately suppressed by $h_0$, as $L_Z/h_0$ increases further, however, because the thermocline suppresses the downward transport of heat. The scatter of data at larger $L_Z/h_0$ may reflect that $h$ can be affected by other factors such as stratification in the residual layer.

In conclusion, the depth of a diurnal thermocline $h$ should be scaled by $L_Z$, instead of $L_{MO}$, contrary to PGPB. We also showed that $h$ is not affected by $h_0$ as long as $h_0$ is sufficiently larger than $h^*$, contrary to the argument by PGPB. Finally, we should mention that the simulation with $L_a = 0.3$, the same as in PGPB, produced essentially the same results; for example, all values of $h$ in Fig. 1 remain the same.

Acknowledgments. This work was supported by the National Research Foundation of Korea grant funded by the Korean government (MEST Grant NRF-2009-C1AAA001-0093068).

REFERENCES


